Response to reviewer 1

The reviewer's comments are in black and our answers are in red.
Modifications of the manuscript are reported in bold and italic.
The pages and lines reported here correspond to the original pdf.
New references can be found at the end of the document.

General comments:

In this paper, the authors present temporal evolution of summer time near surface snow specific surface area (SSA) at Dome C, Antarctica estimated from two types of in-situ measurements (performed during two summer campaigns: 2012–2013 and 2013–2014), and satellite remote sensing in the microwave region (obtained during 2000–2014). In addition, they investigate whether the Crocus snowpack model forced by ERA-Interim reanalysis data can be used as a useful tool to understand observed changes in near surface SSA. In conclusion, they state that observed variations of near surface SSA were successfully reproduced by Crocus; however, effects of wind on the snow compaction and SSA evolution can be overestimated in the model. Overall, this manuscript is well written and easy to follow. Provided information are valuable for TC readers who are interested in not only physics of SSA but also surface energy balance in Antarctica. In addition, snow modelers might also find this manuscript interesting. Therefore, this reviewer recommends its publication once the authors attend to the following comments. My major concern is whether the authors have confirmed the adequacy of ERA-Interim data (note that this is not observation) in Antarctica. If the accuracy of input data for Crocus (ERA-Interim) is insufficient, the reliability of presented model performance in this study can be somewhat lowered. In the following part, this reviewer gave specific comments. Please note that page and line numbers are denoted by “P” and “L”, respectively.

We are grateful to this reviewer for the encouraging comments, and tried to emphasize the adequacy of ERA-Interim data. This point has been addressed in more details by Libois et al. (2014a). Here, the paragraph dedicated to ERA-Interim data has been largely expanded as explained below in our response to the specific comments.

Specific comments:

P4501, L7-9: SSA also controls the e-folding depth as well.
This was added in the text:
Snow specific surface area [...] ***strongly affects*** snow albedo ***and light e-folding depth***, especially [...].

P4502, L22: Please describe more why detailed snowpack models are not fully adequate for polar environments.
We added the following sentence:
“***In fact,*** such models are usually not fully adequate to polar environments (Dang et al., 1997, Groot Zwaaftink et al., 2013). ***Their semi-empirical parameterizations for snow metamorphism, compaction and fresh snow characteristics are indeed often based on observations made in alpine environments (e.g. Marbouty, 1980, Guyomarc'h et al., 1998), and do not necessarily perform well in colder and drier areas. In addition,***”

P4505, L3-5: Does it mean that the authors used data obtained only under clear-sky conditions? Please clarify.
The method based on the clear-sky parameterization was applied to all data independently of the sky

conditions. It means that cloudy conditions are included in the retrieved SSA time series. This point is now clearly specified:

P4505, L2:
"Since this information is not available from measurements, the direct/diffuse ratio was supposed to depend only on SZA, ***and was thus treated in the same way for clear-sky and cloudy conditions***."

P4506, L17:
"This procedure is applied ***every day independently of sky conditions. It is repeated every year from 18 October to 27 February***, when SZA at noon remains lower than 67°."

P4509, L1-6: Please indicate expected accuracy of this remote sensing technique.

The accuracy of this remote sensing technique was not mentioned by Picard et al. (2012) and has not been addressed since then. However, it is likely that the largest source of uncertainty is the density value chosen in the inversion procedure. Here snow density was assumed equal to 320 kg m^{-3}, which corresponds to the average value observed at Dome C, but we tested the inversion for 300 and 350 kg m^{-3} as well, which essentially corresponds to the variability observed in the field. We believe that the obtained SSA range gives an estimate of the accuracy of the method. Although it depends on SSA, it is estimated to be roughly 40%. This information was added in the text as follows p4509, L4:

"As a result, this SSA time-series is not expected to be as accurate as the spectrometry-based approach described in Sect. 2.1.1. ***The most critical assumption is probably that of constant density. Assuming a density of 300 kg m^{-3} (respectively 350 kg m^{-3}) instead of 320 kg m^{-3} yields SSA differences up to +20% (respectively -40%), which gives a broad estimate of 40% for the accuracy of the method***."

P4509, L13: This reviewer could not understand why the authors listed "sphericity" here. This is a "virtual" parameter.
In fact the listed variables are ***prognostic*** variables (not diagnostic, updated P4509, L12). Although sphericity may be considered a variable with loose physical definition, it evolves from a time step to another and is used to compute snow compaction by the wind for instance. Here it is mentioned to point out the difference with the former version of Crocus (Brun et al., 1989, 1992) which also included the ad-hoc variables dendricity and grain size (replaced by SSA by Carmagnola et al., 2014). Hence sphericity was maintained in the list of variables.

P4510, L3-6: The explanation provided here is a bit difficult to follow. Please describe in more detail.
This explanation was reformulated to be more understandable P4510, L1:
"***SSA decrease is computed from the formulation F06 of snow metamorphism (Carmagnola et al., 2014) which is based on a fit of the semi-empirical microphysical model of SSA decrease rate proposed by Flanner et al. (2006). Because of working in the mid-latitude context, the fit in Carmagnola et al., 2014 was computed over a period of 14 days, as in Oleson et al. (2010). Here we use the same approach but extend the period to*** 100 days to account for the slower metamorphism resulting from the low temperatures prevailing at Dome C."

P4510, L24: typo: "both were both . . ."
corrected

P4510, L27: The authors introduce ERA-Interim to drive Crocus in this study. Have the authors confirmed its accuracy in Antarctica? In case systematic biases were found in some properties, did the authors correct them?
ERA-Interim data accuracy was investigated by several authors and more specifically to run Crocus in Libois et al. (2014a) and Fréville et al. (2014). Only the precipitation rate was corrected because it showed a significant negative bias at Dome C. Based on in situ observations of 10 m wind and 2 m air temperature, the latter ERA-Interim data seem adequate at Dome C. At least they do not show any significant bias and were thus used as is. This is now detailed as follows:

"Crocus was forced by ***3-hourly*** ERA-Interim atmospheric reanalysis for 2 m air temperature and specific humidity, surface pressure, precipitation amount, 10 m wind speed, and downward radiative fluxes. ***ERA-Interim data were already used by Fréville et al. (2014) to simulate snow surface temperature on the Antarctic Plateau. As detailed in Libois et al. (2014a), precipitation rate was multiplied by 1.5 to ensure that simulated annual snow accumulation matches observations at Dome C. On the contrary, ERA-Interim wind was found in good agreement with measurements performed on the 40 m high instrumented tower at Dome C (Genthon et al., 2013). Libois et al. (2014a) also pointed that drift events observed at Dome C could satisfactorily be predicted from ERA-Interim wind time series, further supporting the consistency of wind data. As for air temperature, it does not show any significant bias during the summer from 2000 to 2013 compared to Dome C II automatic weather station (http://amrc.ssec.wis.edu/aws). It does show a positive bias of about 2K during the winter, but this is not critical for our study because snow metamorphism barely operates in winter.*** "

P4510, L27 – P4511, L1: Please indicate time intervals of ERA-Interim and output data from Crocus simulations.
ERA-Interim data were prepared at 3-hourly time step by composing analysis and short term forecasts. This was added (see previous comment).
Crocus is run at a 15 min time step but output data for this study are considered every 12 hours resolution because our main focus is on snow metamorphism, which operates at the scale of several days.

P4511, L4:
"Then, Crocus was run from 2000 to 2014 ***and the full state of the snowpack was recorded every 12h,*** yielding the reference simulation that is analysed in the following."

P4511, L1: Please indicate how many model layers were set in the 12 m snowpack. In addition, it might be informative to list model layer thicknesses set in this study.
This information was added as follows P4511, L11:
"The snowpack was first initialized with a~depth of 12 m [...]. ***It comprised 25 layers***."

We also precised that the number of snow layers is variable in Crocus in case it was not clear for the readers (P4509, L11):
"The number and thickness of numerical snow layers evolve with time."

and P4509, L13:
"Crocus was adapted to the specific meteorological conditions prevailing at Dome C [...]. ***In particular, the optimal thicknesses of the 5 topmost layers were set at 2, 3, 5, 5 and 10 mm, to ensure that surface processes are accurately represented***."

P4512, L9: It seems to me that the title of Sect. 3.1 "Daily variations of SSA" is not suitable, because

data intervals presented in Fig. 3 are several days (not a few hours or less).
The title was changed into "***Seasonal variations of SSA in the uppermost 2 mm***"
To maintain consistency from a section to another. The titles of Sect. 3.2 and 3.3 were also changed:
"Seasonal variations of SSA ***in the uppermost 2 and 10 cm***"
"Inter-annual variability ***of SSA in the uppermost 10 cm***"

P4513, L14: Does this explanation mean that the top most model layer thickness of Crocus is less than 2 mm?
The thickness of Crocus layers evolve depending on compaction, precipitation, sublimation, etc. Layers can also be merged or split depending on their properties. The model always tries to match an optimal thickness profile. Here this optimal profile is 2, 3, 5, 5 and 10 mm for the uppermost 5 layers as mentioned above. It means that the uppermost layer thickness tends to be around 2 mm. After a light precipitation event, the topmost layer can be 1 mm thick. In this case, the computation over the topmost 2 mm implies that at least 2 layers are accounted for. In the case the topmost layer is more than 2 mm, the average SSA over the topmost 2 mm is simply that of this layer. The information regarding layer thickness has been added before.

P4513, L23-26: Please discuss why Crocus could not simulate the effect of soft snow removal by the wind, and the formation of surface hoar.
Such processes are indeed currently not simulated explicitly by Crocus. The last sentence of the conclusion clearly states that these processes could be simulated by Crocus and should be regarded as potential improvements.

P4521, L8:
"Other physical processes not yet simulated by Crocus should also be regarded as potential progress for simulating snow properties on the Antarctic Plateau, such as the formation of hoar crystals, and the mixing of the topmost layers of the snowpack due to snow drift."

Currently, the only way Crocus can loose mass to the atmosphere is via sublimation, which in the model does not lead to any change in snow physical properties. The effect of the wind is essentially to compact snow and possibly to increase SSA as a result of smaller ice crystals falling last after a drift event. In the model, wind does not physically remove snow, it can only foster sublimation, which is a very different process than that observed at Dome C. As for surface hoar, condensation can occur on top of the snowpack, but the newly deposited snow has the same properties (density and SSA) as the uppermost layer, which is quite different from what is observed in the field and called surface hoar. In addition, surface hoar resulting from vapor transfers within the snowpack is not simulated at all because such mass transfer is not simulated so far, although work is in progress on this critical question.

To make it clear we slightly modified the sentence P4513, L23:
"The effect of soft snow removal by the wind as well as the formation of surface hoar are ***currently*** not simulated by Crocus [...]."

P4514, L20-25: This reviewer could not follow what the authors intended to explain here. It might be better to reformulate.
There are 2 ways to compute the surface SSA. Either to compute a linear average, or an exponential decay "average". The latter accounts for the fact that the uppermost layers (few mm) contribute more to the albedo than layers below. The exponential decay correspond to the light e-folding depth in snow. The paragraph was reformulated as follows:

"***Since solar irradiance decreases exponentially with depth, the uppermost mm of the snowpack contribute more to the albedo than the snow below. As a result, the SSA retrieved from albedo measurements is the result of a convolution of the actual SSA profile by an exponential to a first approximation. To account for this effect, the simulated SSA was also computed using*** a 2 cm exponential decay (Mary et al., 2013) rather than a linear average. ***This resulted in slightly higher SSA (less than 5%). Likewise, since the choice of 2 cm is to some extent arbitrary, the average was also computed over the topmost 1 and 4 cm***. ***It resulted*** in less than [...]"

P4515, L1-2: The contrasting feature of summer SSA decrease between 2012–2013 and 2013–2014 is interesting. Could the authors discuss the reason of this difference by referring to meteorological conditions during these two summers?
The analysis of ERA-Interim precipitation shows that during the winter 2012, the total amount of precipitation has been 45% more than during the winter 2013. This information was added in the discussion:

"The summer decrease was thus more significant in 2012-2013 than in 2013-2014, which is reproduced by Crocus (Fig. 5). ***More precisely, the main difference between both summers is the initial value of SSA. This can be explained by the fact that ERA-Interim precipitation accumulated from March 1st to November 1st was 45% larger in 2012 than in 2013***."

P4520, L6-15: Before discussing the impact of wind speed on the topmost 7 cm SSA evolution, the authors should demonstrate accuracy of wind speed obtained from ERA-Interim (related to "P4510, L27"). If wind speed from ERA-Interim is overestimated, this discussion has no meaning.
We believe that wind speed at 10 m is sufficiently well simulated by ERA-Interim, and does not show any particular bias. This was detailed previously (response to "P4510, L27"), and is much more detailed in the study by Libois et al. (2014a) dedicated to the impact of wind on snow properties.

P4520, L20: The validity of meteorological forcing used in this study has not been confirmed (related to "P4510, L27").
See previous comment and "P4510, L27".

Figure 1: "mat": typo?
Corrected

Figure 3a: Two hatched areas are difficult to distinguish from each other.
We replaced the hatched areas by shaded areas .

Figure 3b: What do the authors mean by the "dark line"? In addition, what are the dark dots? This is not explained in the caption.
In fact what you call dark dots were referred in the caption as "clear dots" and the dark line referred to the line linking the white dots. This was indeed misleading, and was clarified in the caption as follows:

"**The grey circles indicate single measurements and the white circles highlight the median value for each day**."

References:

Genthon, C., Six, D., Gallée, H., Grigioni, P., & Pellegrini, A. (2013). Two years of atmospheric boundary layer observations on a 45-m tower at Dome C on the Antarctic plateau. *Journal of Geophysical Research: Atmospheres*, *118*(8), 3218-3232.

Guyomarc'h, G.., & Mérindol, L. (1998). Validation of an application for forecasting blowing snow. *Annals of Glaciology*, *26*, 138-143.

Marbouty, D. (1980). An experimental study of temperature gradient metamorphism. *Journal of Glaciology, 26*, 303-312.

Response to reviewer 2

The reviewer's comments are in black and our answers are in red.
Modifications of the manuscript are reported in bold and italic.
The pages and lines reported here correspond to the original pdf.
New references can be found at the end of the document.

General comments:

This study addresses the seasonal and inter-annual variability of the near-surface specific surface area (SSA) at Dome C, on the Antarctic Plateau. SSA is derived from optical measurements, from high frequency microwave observations (89, 150 GHz), and is simulated with the snow model Crocus, which is forced by atmospheric quantities from ERA-Interim reanalysis. The topic is certainly very relevant, as the understanding and quantification of the SSA evolution allow the understanding, quantification, and better simulation of the snow mass and surface energy budgets. The SSA derived from observations is used to test the Crocus capability to simulate the various physical processes contributing to the daily, seasonal, and inter-annual variability of SSA. The paper is generally well written and well argued. The weakest point is the insufficient error analysis of the spectral albedo measurements and of the SSA derived from them. The SSA derived from albedo measurements was found to be in very good agreement with the SSA simulated with Crocus. However, a better error analysis of the measurements would strengthen the Crocus validation, as the Crocus simulations are in any case based on parameters specifically adjusted to the Antarctic environment. Also the uncertainties on the SSA derived from microwave measurements could be better assessed. In conclusion, I consider the paper well suited for publication in The Cryosphere after a minor revision, which can be done addressing the specific comment listed below.

We thank the reviewer for the attention paid to our study and modified the manuscript to better account for measurements errors. Here, we do not pretend to develop a thorough error analysis of the albedo measurement and derived SSA, because such an exercise requires a dedicated study. The latter is work in progress and should be submitted soon. We have however added new elements that enable us to derive an estimation of the albedo error and retrieved SSA. It is based on a better analysis of the scaling coefficient A, whose meaning was more detailed according to the reviewer's recommendations. We also stress that our main interest is in SSA variations rather than SSA absolute values. All these points are addressed in details in the specific comments.

Specific comments:

p.4501, lines 3-5: “SSA determines the albedo, especially in the near-infrared” is quite a rough statement. Although SSA has a first order impact on albedo, it does not entirely determined the albedo, as snow density, snow particle shape, and other microstructural characteristics have a second order impact on albedo. Perhaps instead of using “determine”, the authors can write that “SSA controls”, or “strongly affects” the albedo. In fact, in the following sentence the authors write “especially”, contradicting the statement that albedo is solely determined by SSA.
We used “***strongly affects***” instead of “determines”.

p.4502, line 7-9: “when solar energy is absorbed deeper, it warms up the snowpack and increases temperature gradients, which in turn enhances metamorphism close to the surface and e-folding depth”. I would remove from the sentence “and e-folding depth”, as it is not certain that the e-folding depth increases when the surface layer becomes more absorptive due to the metamorphism (and for instance snow crusts form).

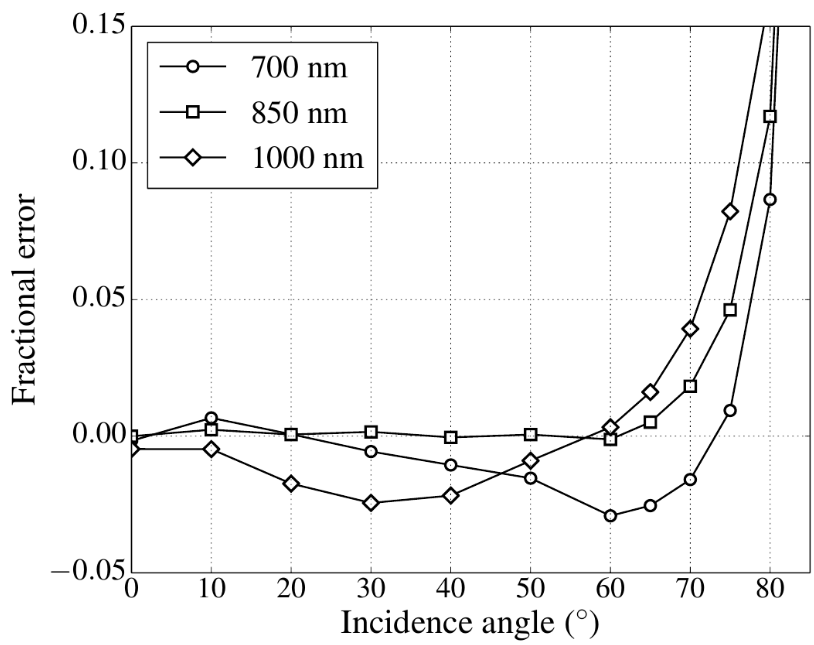
In fact the feedback loop mentioned here requires the e-folding depth to increase with metamorphism. Otherwise, only the more usual albedo feedback should be invoked. The formation of snow crusts as a result of dry metamorphism is not well understood at Dome C and there is no evidence to our knowledge that it leads to more or less absorptive layers. Here metamorphism implies SSA decrease, so this detail was added to avoid erroneous interpretation:

"[...] it warms up the snowpack and increases temperature gradients, which in turn enhances metamorphism close to the surface. ***As a consequence, SSA generally decreases and e-folding depth increases.***"

pp. 4504-4505, Sect. 2.1.1. As the instrument to measure spectral albedo is newly designed, it would be good to better quantify its accuracy, especially with respect to the deviation from the ideal cosine response (a plot with the deviation of the ideal cosine response as a function of incident angle would be welcomed, instead of just mentioning that the angular response was determined in the laboratory).

The detailed analysis of the instrument accuracy will be presented in another study to be submitted in a few weeks. For the present study, more details are however added.

The figure below shows the deviation of our collectors response from the ideal cosine response as a function of incident angle for 700, 850 and 1000 nm. This is similar to Fig. 2 of Grenfell et al. (1994):



We do not think that the plot is worth being added to the manuscript because the detailed analysis of the collector response will be investigated in the furure study. However, the quantitative information about the collector response was added as follows P4504, L26:

"To this end, the angular response of our collectors was determined in the laboratory. ***The deviation from the perfect cosine response is less than 4% for angles below 70°, but increases beyond 80° due to the dome geometry of our collectors that capture a significant amount of light at grazing angles (Bernhard et al., 1997).***"

The upward and downward looking fiber optics require two specific irradiance calibrations, which include some inaccuracy. Can the authors exclude an even small systematic error in the albedo calculated from the ratio between the two signals?

Although we use the same angular correction for all the collectors because they were all designed in the same way, it is necessary to inter-calibrate the upward and downward looking fibers because overall transmittance of the optical system is not the same for both lines. To do this, the two collectors of the same measurement head were consecutively positioned in the downward-looking position, by simply flipping the arm (Fig. 1) of 180°. Horizontality was carefully checked for this procedure and we assumed the upward flux remained constant during this ~1 min experiment. The ratio of the spectra obtained with both lines was used to rescale the albedos analysed in the study. This inter-calibration step is now detailed in the text P4505, L3:

"It was calculated at all wavelengths with the atmospheric radiative transfer model SBDART (Richiazzi et al., 1998) for typical summer clear-sky conditions at Dome C. ***Although the upward- and downward-looking fibers are assumed to have the same angular response, the transmittances of both optical lines are different. To account for this effect, both lines were inter-calibrated. For this, the two collectors were set in the downward-looking position, by simply flipping vertically the fibers of 180°. The procedure lasted less than 30 s and was performed on a clear-sky day, so that the upward flux could be assumed constant. The ratio of both spectra was used to rescale all the albedos analysed in this study***."

What about the horizontal levelling of the cosine collectors, was it regularly checked? A misalignment of few degrees could well explain the observed excessively high albedo in the visible.

The system horizontality was checked at installation with embedded electronic sensors. At this time it was 0.4° and is accounted for in the procedure to correct the incident radiation. Unfortunately the inclinometer did not work in the cold and no continuous measurement was available. It has not been measured since then. However, the horizontality was estimated from the symmetry of the incident radiation around local noon. There is so far no reason to believe that a significant tilt is present in the system. The manuscript was modified as follows:

P4504, L12:
"It has 2 similar measurement heads looking to the surface and to the sky (Fig. 1). ***The horizontality of the heads was checked at installation with an electronic inclinometer and was better than 0.5°***."

Was the impact of the shadow of the whole measuring system accounted for in the albedo calculation?

The shadow of the instrument is not accounted for, essentially because measurements are taken at noon and at that time the thin shadow of the mast is far from the head (Fig. 1) and its impact likely to be minor. It is now detailed in Fig. 1 caption that the picture was taken at ***11:00 local time***. In addition, the question of this shadow is mentioned in the discussion about the scaling coefficient A below.

When the authors applied the correction for the angular response following the method of Grenfell et al. (1994) did they assume isotropic reflection from the snow surface? Also this assumption can be the source of a small error (see Carmagnola, C. M., Dominé, F., Dumont, M., Wright, P., Strellis, B., Bergin, M., Dibb, J., Picard, G., Libois, Q., Arnaud, L., and Morin, S.: Snow spectral albedo at Summit, Greenland: measurements and numerical simulations based on physical and chemical properties of the snowpack, The Cryosphere, 7, 1139-1160, doi:10.5194/tc-7-1139-2013, 2013).

Indeed, we assumed the reflected radiation was isotropic. According to Carmagnola et al., (2013), this assumption results in an error of 0.2-0.4%, and again is added in the factors affecting the value of the scaling coefficient A. It is now mentioned explicitly P4505, L3:

"It was calculated at all wavelengths with the atmospheric radiative transfer model SBDART (Richiazzi et al., 1998) for typical summer clear-sky conditions at Dome C. ***Contrary to incident radiation, reflected radiation is assumed isotropic."***

In conclusion, it would be important to estimate the error in the albedo that remains after the applied correction of the angular response (following the method of Grenfell et al., 1994), and calculate how this error propagates to the estimated SSA.

The remaining error after all corrections are applied depend on : the uncertainty on collector response measurement, the variability of direct/diffuse irradiance, the instrument horizontality, the effect of snow anisotropy, the assumption of surface flatness, the impact of shadowing... Since the objective of the present study is not to derive an accurate measurement error using forward error modelling, we use an empirical approach to get an estimate of the precision. In fact, the scaling coefficient A somehow includes all the remaining errors, and should equal 1 if measurements were perfect. Practically the time series of the coefficient A shows that it essentially varies from 0.98 to 1.03. Hence an upper bound error is chosen for the estimated precision of albedo measurements: 3%. To convert this albedo error into SSA error, we use Eq. (1). For the spectral range and SSA of interest, the accuracy is expected to be better than 25%. This information was added in the manuscript P4506, L23:

"The SSA retrieved with this algorithm roughly corresponds to the SSA of the top 2 cm of the snowpack [...]. ***A rigorous forward estimation of the accuracy of the algorithm would require a thorough analysis of several factors including the uncertainty on the collector calibration procedure, the effect of snow anisotropy (Carmagnola et al., 2013), the shadowing of the surface by the instrument, the potential tilt of the sensor, the validity of the semi-infinite snowpack assumption, etc. Taking into account all these factors and their inter-correlation is beyond the scope of this article and will be addressed in future work. Here the accuracy is estimated using a global approach based on the analysis of the coefficient A obtained during the retrieval. Over the period of observations, it varied in the range 0.98 - 1.03, while ideal measurements would have yielded A=1. The deviation of A from 1, that is -2% to +3%, gives an estimation of the albedo measurement accuracy. Hence we assume that the albedo accuracy is 3%. The corresponding accuracy on SSA estimation is then derived from Eq. 1. For the spectral range of interest and the SSA values encountered at Dome C, the estimated accuracy of the SSA retrieval is better than 25%***."

P 4505, line 23: For Eq. (2) the mentioned reference is not correct. A correct reference is for instance Negi, H. S. and Kokhanovsky, A.: Retrieval of snow albedo and grain size using reflectance measurements in Himalayan basin, The Cryosphere, 5, 203–
217, doi:10.5194/tc-5-203-2011, 2011.

Reference to Negi et al. (2011) was added here.

p.4506, lines 10-14: the authors introduce the coefficient "A" to deal with the uncertainty on albedo measurements. However, the definition and calculation of the coefficient "A" is quite confusing: is "A" wavelength dependent? The main problem here is the lack of a proper characterization and quantification of the errors in the albedo measurements. If the albedo cannot be further corrected, and the remaining error is partly attributable to the deviation from an ideal cosine response of the instrument, then the error in the albedo is wavelength dependent, as generally the deviation of the cosine response given by diffusers is wavelength dependent. Thus, a constant “A” through the analysed wavelength range introduces an artefact. On the other hand, if “A” is wavelength dependent, it cannot be uniquely determined together with SSA using solely Eq.(3). Finally, if “A” is related to the error in the measured albedo, its value should be shown and commented (although, I think that the error quantification should be more directly and clearly expressed than through the coefficient “A”).

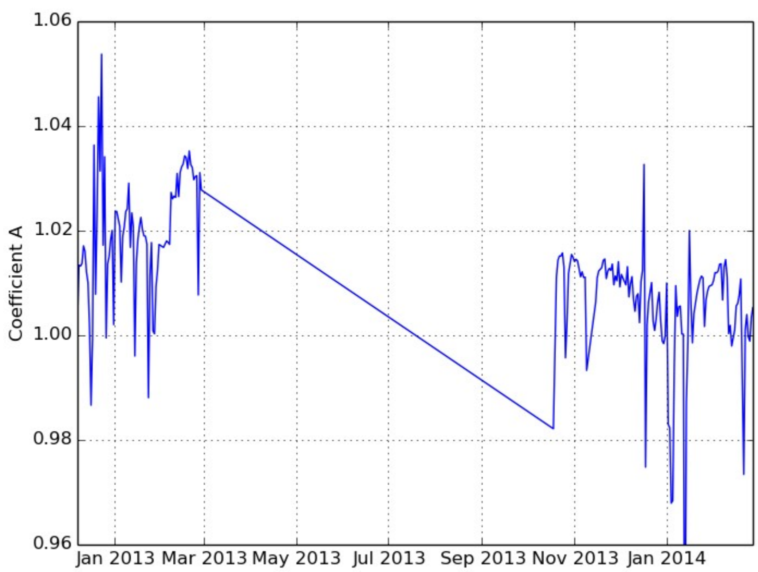
We clarified the definition of A. In fact, A somehow includes all the errors that were not accounted for before. A is indeed wavelength independent here. We believe that A mainly results from geometrical errors that are less wavelength-dependent than the deviation of the cosine response, the latter already being corrected. The text was modified P4506, L10:

“***To account for remaining uncertainties in the albedo spectra, a scaling coefficient A*** is optimized along with SSA, so that the optimized function α is actually given by:

Eq. (3)

where r_λ^{diff} gives the proportion of diffuse light. ***A is meant to compensate for all the factors affecting in a wavelength-independent way albedo measurements, that were not explicity corrected by the previous processing steps.***“

The evolution of A for the two seasons 2012-2013 and 2013-2014 is shown below. However, we think this figure is unncessary in the paper.



As it seems unclear to the reviewer that the deviation from the ideal cosine is already accounted for, it is now more clearly stated:

P4504, L25:
“Despite our effort to build highly diffusing collectors, ***small remaining deviations from the ideal cosine response need to be corrected***.”

P4504, L26:
"To this end, the angular response of our collectors was determined in the laboratory, ***and used to estimate the true incident and reflected fluxes from the measured ones (Grenfell et al., 1994).***"

P4508, Sect 2.2: what is the estimated accuracy (from literature) of the ERA-Interim air temperature used as input to the DMRT-ML model to simulate the SSA? And what is the variability (std) of the snow density during the summertime? The air temperature accuracy and the snow density variability could be used to assess the sensitivity of the retrieved SSA to these uncertainties.

Although recent studies have shown a warm bias of ERA-Interim temperature on the Antarctic Plateau (e.g. Fréville et al., 2014), we are not aware of any thorough evaluation of ERA Interim temperatures at Dome C. On the contrary, we have performed many measurements of snow density at Dome C, and this parameter essentially varies between 300 and 350 kg m^{-3}. For the inversion it was assumed equal to 320 kg m^{-3} which corresponds to the average value observed, but the inversion was applied for 300 and 350 kg m^{-3} as well. The obtained variability gives an estimate of the accuracy of the method. Although it depends on SSA, it is estimated to be roughly 40%. This information was added in the text as follows p4509, L4:

"As a result, this SSA time-series is not expected to be as accurate as the spectrometry-based approach described in Sect. 2.1.1. ***The most critical assumption is probably that of constant density. Assuming a density of 300 or 350 kg m^{-3} instead of 320 kg m^{-3} leads to SSA differences of maximum 40%, which we consider to be a rough estimate of the accuracy of the method.***"

We have not estimated the effect of the potential +2K bias of ERA-Interim 2m temperature in winter in our particular case but increasing the snow temperature by 2K corresponds to a variation of emissivity of approximately 0.01 which yields a smaller variation of SSA than the 10% change of density (e.g. Fig. 3, Brucker et al. 2010 for example at 37 GHz).

p. 4509, lines 4-5: here the problem is that the authors have not explained how accurate the spectrometry-based approach to retrieve SSA is.
This accuracy is now explicitly mentioned

p. 4509, line 22: rephrase as "It was reformulated in terms of SSA using Eq. (5) of Carmagnola et al. (2014). . .".
corrected

p. 4510, line 24: "here both were both fixed".
corrected

p. 4512, line 9: rather than "Daily variations of SSA", Section 3.1 describes "Seasonal variations of SSA in the uppermost 2 mm".
Corrected. The titles of Sect. 3.2 and 3.3 were also changed for consistency:
"Seasonal variations of SSA ***in the uppermost 2 and 10 cm***"
"Inter-annual variability ***of SSA in the uppermost 10 cm***"

p.4512, line 10-12: how many SSA values were used in the calculation of the mean SSA for each 1-m transect? Given that the ASSSAP was located 5 cm above the surface, were all the used SSA measurements independent (i.e, was the field of view of the ASSSAP smaller than the distance between

two consecutive measured spots)? If the SSA measurements are not independent, then the standard deviation utilized in Fig 3a to illustrate the SSA variability has a questionable meaning.
ASSSAP measures the snow reflectance every 10 ms. The number of measurements acquired depend on the speed at which the instrument is moved on the horizontal rail by the operator. Practically, the instrument was passed two times along the rail for each transect and about 1000-1500 points were acquired. Afterwards, the measured transect was divided in 1-cm long intervals. The median for all values taken in the same interval was then computed. The dots in Fig. 3a correspond to the mean of these medians. The std shown in Fig. 3a is that of all median values. The footprint of the laser beam on snow is less than 1 cm because light penetration at 1310 nm is only 2 mm, so that the std is a relevant quantity here. The text was modified P4511, L10:

"For each 1 m long horizontal transect taken with ASSSAP in 2012-2013, the average surface SSA in the range 0.25-0.75 m was computed. ***For that, the measured transect was first divided in 1 cm intervals over which the median SSA value was taken. These medians were then used to compute the average value and standard deviation for the transect,*** from which the temporal evolution of SSA at the two locations was deduced (Fig.3a)."

p. 4514, line 1: the title of Section 3.2 could be "Seasonal variations of SSA in the uppermost 2 and 10 cm".
Changed, see p. 4512, line 9 comment

p. 4517, line 26: ". . .one year to another than in Crocus than in the observations".
First "than" removed.

p. 4517, lines 27-29: there is some confusion in explaining SSA evolution in different snow layers and in different time scales. I would rephrase for instance as "Although the impact of snow precipitation seems moderate in Crocus simulations of SSA in the top 2 and 10 cm, snowfall occurrence and amount drive Crocus-simulated SSA variations in the top 2 mm, consistently with observations. While the deeper layers show a seasonal SSA evolution, the surface layer mostly reflects day-to-day SSA variations".
The reviewer's suggestion was incorporated in the text except for the end that wsa replaced by:
"the surface layer mostly reflects day-to-day variations of weather conditions."

p. 4518, line 15: ". . . and makes complicated the comparison between punctual observations and simulations difficult"
"difficult" was removed.

p. 4518, line 21: if the spectral albedo sensors are placed at the height of about 2 m, then 50% (90%) of the received reflected irradiance comes from an area with radius of about 2m (6m) (see Schwerdtfeger, P. (1976), Physical Principles of Micrometeorological Measurements, 113 pp., Elsevier Sci., New York).
This quantitative information and the reference were added as follows:
"Conversely, the spectral albedo measurements cover ***an area with radius of approximately 6 m (Schwerdtfeger, 1976)*** and probe deeper into the snowpack."

P 4518, lines 20-23: The sentences ". . ., which is more likely to be representative of surface snow at Dome C, even though larger-scale spatial variability exists" are quite ambiguous and unclear. It has been explained through the paper that the spectral albedo measurements in the wavelength range 700-1100nm mostly depend on the averaged SSA in the uppermost 2 cm of the snowpack, which also includes the 2-mm-thick surface layer monitored with the ASSAP. If the authors are now comparing the SSA in the two layers (top 2 cm and top 2mm), they cannot state that the former "is more representative of surface snow". What is "surface snow", the top 2cm or the top 2mm? Maybe the authors mean that the SSA derived from the albedo measurements represent a larger area, but of the top 2 cm of snow, not of the very surface (top 2 mm). I would like to remark that, even if albedo was measured at longer wavelengths (1300nm or larger) to get the SSA of the top 2mm from the same large area of ∼6m radius, it not at all sure that the derived SSA would have been in better agreement with the Crocus-modelled SSA. This because the scale of spatial variability of the wind-compacted/eroded and snowdrift-accumulation areas has a quasi-period of 30-50m, as the authors found in another paper (Picard, G., Royer, A., Arnaud, L., and Fily, M.: Influence of meter-scale wind-formed features on the variability of the microwave brightness temperature around Dome C in Antarctica, The Cryosphere, 8, 1105–1119, doi:10.5194/tc-8-1105-2014, 2014). This quasi-period is evidently larger than the footprint of the spectral-albedometer, which then does not necessarily exactly corresponds to the large-scale average snow surface SSA.

The surface snow mentioned here is indeed quite confusing, because it refers to 2 cm, but seems to be compared to that of the transect, ie 2 mm. It has been improved P4518, L21:

"[...] and probe deeper into the snowpack. ***Hence they are more likely to be representative of the average snow SSA in the topmost 2 cm at Dome C***, even though larger-scale spatial variability exists (Picard et al., 2014)."

Fig. 3-6: in all the 4 figures is quite difficult (or impossible) to associate the dates to the plotted data. Perhaps the authors could remove the years from the date labels and mark them as titles of the subplots ("2012-2013" and "2013-2014"). Also, plots could have the grid (horizontal and especially vertical) on.

As recommended by the reviewer, the years are now indicated as titles for each plot, the time labels are now simply 01/12 for January 12th. Grid is on for all Figures 3-6.

p. 4531, line 2 of Figure caption: "mat" should be "mast".
corrected

p. 4533, last line of Figure caption: after "era-Interim" please add "(right y-axis, dark grey columns)".
Done, added for Fig. 4 as well.

p. 4534, Fig. 4: in both subplots, it would be very useful to mark (maybe with a rectangle box?) the section of time series that correspond to Fig. 3. Otherwise, it is difficult to compare Fig.4 with Fig.3.
We've added horizontal double head arrows to highlight the periods corresponding to Fig. 3. It is detailed in the caption as:
"***The horizontal arrows highlight the periods of measurements shown in Fig. 3.***"

References:

Bernhard, G., & Seckmeyer, G. (1997). New entrance optics for solar spectral UV measurements. *Photochemistry and Photobiology*, *65*(6), 923-930.

Brucker, L., Picard, G., & Fily, M. (2010). Snow grain-size profiles deduced from microwave snow emissivities in Antarctica. *Journal of Glaciology*, *56*(197), 514-526.

Negi, H. S., & Kokhanovsky, A. (2011). Retrieval of snow albedo and grain size using reflectance measurements in Himalayan basin. *The Cryosphere*, *5*(1), 203-217.

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Summertime evolution of snow specific surface area close to the surface on the Antarctic Plateau

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Abstract

On the Antarctic Plateau, snow specific surface area (SSA) close to the surface shows complex variations at daily to seasonal scales which affect the surface albedo and in turn the surface energy budget of the ice sheet. While snow metamorphism, precipitation and strong wind events are known to drive SSA variations, usually in opposite ways, their relative contributions remain unclear. Here, a comprehensive set of SSA observations at Dome C is analysed with respect to meteorological conditions to assess the respective roles of these factors. The results show an average two-to-three-fold SSA decrease from October to February in the topmost 10 cm, in response to the increase of air temperature and absorption of solar radiation in the snowpack during spring and summer. Surface SSA is also characterised by significant daily to weekly variations, due to the deposition of small crystals with SSA up to 100 $m^2\,kg^{-1}$ onto the surface during snowfall and blowing snow events. To complement these field observations, the detailed snowpack model Crocus is used to simulate SSA, with the intent to further investigate the previously found correlation between inter-annual variability of summer SSA decrease and summer precipitation amount. To this end, some Crocus parameterizations have been adapted to Dome C conditions, and the model was forced by ERA-Interim reanalysis. It successfully matches the observations at daily to seasonal time scales, except for few cases when snowfalls are not captured by the reanalysis. On the contrary, the inter-annual variability of summer SSA decrease is poorly simulated when compared to 14 years of microwave satellite data sensititve to the near surface SSA. A simulation with disabled summer precipitation confirms the weak influence in the model of the precipitation on metamorphism, with only 6 % enhancement. However we found that disabling strong wind events in the model is sufficient to reconciliate the simulations with the observations. This suggests that Crocus reproduces well the contributions of metamorphism and precipitation on surface SSA, but that snow compaction by the wind might be overestimated in the model.

1 Introduction

The surface energy budget of the Antarctic Plateau depends on snow physical properties (Van As et al., 2005; Brun et al., 2011). Snow specific surface area (the surface area of the ice–air interface per unit mass of snow, hereafter referred as SSA) ~~determines the albedo~~ strongly affects snow albedo and light *e*-folding depth, especially in the near-infrared (NIR, e.g. Domine et al., 2006), and thus controls the amount of solar radiation absorbed by the surface (e.g. Warren, 1982; Gardner and Sharp, 2010; Carmagnola et al., 2013). Snow density controls the light *e*-folding depth ~~in snow~~ (Libois et al., 2013) and the effective thermal conductivity of the snowpack (Sturm et al., 1997; Calonne et al., 2011), among others. Although the surface of the Antarctic Plateau has often been considered homogeneous in space and stable in time, especially for the calibration of satellite radiometers (e.g. Loeb, 1997; Six et al., 2004), recent studies pointed out that it is subject to large and rapid variations (e.g. Bindschadler et al., 2005; Lacroix et al., 2009; Champollion et al., 2013). Snow properties evolve over time in response to internal thermodynamical processes such as snow metamorphism (Colbeck, 1983) and densification (Gallée et al., 2001). The surface is also affected by meteorological events such as snowfall and wind events (Kuhn et al., 1977; Champollion et al., 2013). The resulting dependence between snow physical properties and the energy budget of the snowpack gives rise to feedbacks (e.g. Albert et al., 2004) and is therefore of great interest for climate studies. This highlights the need to identify the main processes which drive surface snow evolution on the Antarctic Plateau.

The Plateau is characterized by very low temperatures (annual average around $-50\,^{\circ}$C, Augustin et al., 2004), low precipitable water vapour content (less than 1 mm, Tremblin et al., 2011) and low annual accumulation (less than 30 $\mathrm{kg\,m^{-2}}$, Frezzotti et al., 2004), so that the physical processes controlling snow characteristics are substantially different from other environments (e.g. Alps, tundra, ...). During the polar night, temperatures usually remain well below $-50\,^{\circ}$C and snow metamorphism barely operates (Town et al., 2008). On the contrary, at the end of spring, the increase of snow temperature causes significant metamorphism, which leads to an overall decrease of snow SSA (e.g. Picard et al., 2012), densification (Fujita et al., 2009) and other morphological changes of the surface snow (Gow, 1965). As a result, albedo decreases by several percents (Jin et al., 2008; Wang and Zender, 2011), which significantly alters the surface energy budget of the snowpack (van den Broeke, 2004). Picard et al. (2012) have shown that this interdependence between snow optical properties and SSA accelerates snow metamorphism through a positive feedback: when solar energy is absorbed deeper, it warms up the snowpack and increases temperature gradients, which in turn enhances metamorphism close to the surface. As a consequence, SSA generally decreases and *e*-folding depth increases. This positive feedback involving the e-folding depth adds up to the snow albedo feedback (e.g. Flanner and Zender, 2006; Box et al., 2012), making summer metamorphism very sensitive to snow optical properties at the surface.

Hitherto, this summertime SSA decrease has been generally deduced from albedo measurements (Jin et al., 2008; Kuipers Munneke et al., 2008), which also depend on illumination conditions, cloudiness and surface roughness (Wang and Zender, 2011), but such decrease has seldom been measured directly in the field. In addition, the inter-annual variability of summer metamorphism is poorly understood. Picard et al. (2012) showed a strong correlation with summer amount of precipitation, and hypothesized a strong inhibition of the above mentioned positive feedbacks. They used a simple snow evolution model to support their hypothesis, but these features of the seasonal cycle of SSA have never been simulated with a more detailed snowpack model ~~. First because~~ such as Crocus. In fact, such models are usually not fully adequate to polar environments (Dang et al., 1997; Groot Zwaaftink et al., 2013)~~, second because~~ . Their semi-empirical parameterizations for e.g. snow metamorphism, compaction and fresh snow characteristics are indeed often based on observations made in alpine environments (e.g. Marbouty, 1980; Guyomarc'h et al., 1998) , and do not necessarily perform well in colder an drier areas. In addition, the few studies dedicated to the simulation of snow physical properties on the Antarctic Plateau focus on the evolution of the snowpack internal and surface temperatures (e.g. Brun et al., 2011; Fréville et al., 2014), or on punctual profiles (Dang et al., 1997; Groot Zwaaftink et al., 2013), rather than on temporal evolution of snow properties. Nevertheless, correctly simulating SSA evolution remains crucial to better understand the sensitivity of this region to future changes in precipitation and air temperature (Krinner et al., 2006).

The aim of this paper is to investigate the summertime evolution of snow SSA at Dome C, and to further understand its variability, from the daily to the inter-annual scale. To quantify this evolution, we use three datasets. Firstly, a large number of in situ SSA measurements were collected at Dome C during summer campaigns (Sect. 2.1). These included vertical profiles between the surface and 10 cm, and snow samples from the surface, both measured manually during the summer campaigns 2012–2013 and 2013–2014 using the SSA profiler ASSSAP (a light version of POSSSUM, Arnaud et al., 2011). In addition, automatic measurements of snow spectral albedo were used to estimate the evolution of SSA close to the surface during daylight periods. Secondly, the evolution of SSA in the topmost centimetres was estimated from remote sensing observations of the snowpack in the microwave range over the period 2000–2014 based on Picard et al. (2012) (Sect. 2.2). Thirdly, the detailed snowpack model SURFEX/ISBA-Crocus (hereinafter referred as Crocus, Brun et al., 1989, 1992; Vionnet et al., 2012) was used to simulate snow SSA at Dome C (Sect. 2.3). For this, it was adapted to Dome C conditions by changing some parameterizations (Libois et al., 2014a), and forced by ERA-Interim near-surface reanalysis (Dee et al., 2011). The observations and simulations are compared in Sect. 3. The inter-annual variability of summer metamorphism is eventually investigated in more details with Crocus. In particular, the sensitivity of simulated SSA to changes in precipitation, wind and temperature are estimated, which also helps identifying the current potential and limits of Crocus (Sect. 4).

2 Materials and methods

The temporal variations of snow SSA at Dome C were estimated from in situ measurements and satellite data. Snow spectral albedo in the visible and NIR range has been measured using a specifically designed automatic instrument, from which surface SSA variations were deduced over the summers 2012–2013 and 2013–2014. To complement these automatic measurements and to explore the variations deeper in the snowpack, supplementary SSA measurements were taken manually with the instrument ASSSAP. These comprised surface SSA measurements, as well as vertical profiles down to 10 cm depth. Eventually, to study the inter-annual variations of surface SSA, satellite measurements of microwave brightness temperature were used.

2.1 Field observations

2.1.1 SSA estimation from spectral albedo measurements

Using the dependence between snow albedo and SSA close to the surface (Warren, 1982; Domine et al., 2006), variations of the latter were estimated from spectral albedo measurements in the range 300–1100 nm (3 nm resolution). Albedo was recorded every 12 min at approximately 600 m West of Concordia station (75.1° S, 123.3° E, 3230 m a.s.l.), in the "clean area" (75.09960° S, 123.30244° E). The instrument was deployed on 10 December 2012 and has been running almost continuously since then. It has 2 similar measurement heads looking to the surface and to the sky (Fig. 1). ~~One~~ The horizontality of the heads was checked at installation with an electronic inclinometer and was better than 0.5°. One head is optimized for measurements in the UV and visible, the other for visible and NIR. Only the data from the visible-NIR head were used in the present study. Each measurement head consists of one upward- and one downward-looking fiber optics mounted with specially designed cosine collectors having a 180° field of view. The fibers are sequentially connected to an Ocean Optics® Maya Pro spectrophotometer through an optical multiplexer. The spectral albedo is obtained by computing the ratio of upwelling to downwelling irradiance after calibration of the raw measurements. We developed an algorithm to estimate the time series of snow SSA from these measurements. Its steps are computed as follows:

1. *Albedo correction.* Downwelling hemispherical irradiance measurements are strongly affected by the quality of the cosine response of the light collector, especially at high solar zenith angles (SZA) typical of the Antarctic Plateau (e.g. van den Broeke, 2004). Despite our effort to build highly diffusing collectors, small remaining ~~errors~~ deviations from the ideal cosine response need to be corrected. To this end, the angular re-

sponse of our collectors was determined in the laboratory, and used to estimate the true incident and reflected fluxes from the measured ones (Grenfell et al., 1994) . The deviation from the perfect cosine response is less than 4 % for angles below 70°, but increases beyond 80° due to the dome geometry of our collectors that capture a significant amount of light at grazing angles (e.g. Bernhard et al., 1997) . In addition, the correction requires knowledge of the direct vs. diffuse parts of the incident flux (Grenfell et al., 1994). Since this information is not available from measurements, the direct/diffuse ratio was supposed to depend only on SZA, and was thus treated in the same way for clear-sky and cloudy conditions. It was calculated at all wavelengths with the atmospheric radiative transfer model SBDART (Ricchiazzi et al., 1998) for typical summer clear-sky conditions at Dome C. ~~Despite~~ Contrary to incident radiation, reflected radiation is assumed isotropic. Although the upward- and downward-looking fibers are assumed to have the same angular response, the transmittances of both optical lines are different. To account for this effect, both lines were inter-calibrated. For this, the two collectors were consecutively set in the downward-looking position, by simply flipping vertically the fibers of 180°. The procedure lasted less than 30 s and was performed on a clear-sky day, so that the upward flux could be assumed constant. The ratio of both spectra was used to rescale all the albedos analysed in this study. Despite all these precautions, albedo values sometimes exceed 1.0 in the visible range at high SZA and occasionally reach up to 1.05, indicating insufficient correction.

2. *Daily computation of albedo.* To minimize the effect described above, we consider only albedo at noon. For this, the 5 measurements taken between 11:30 and 12:30 LT are averaged every day. This means that albedo measurements are taken at constant solar azimuth angle throughout the summer, but not necessarily constant SZA. This choice is also made because preferential orientation of surface relief is known to translate into an azimuthal dependence of albedo (Wang and Zender, 2011), an artefact that should be avoided here. We nevertheless tried to use constant SZA (i.e. variable

local hour and azimuth) which minimizes errors due to imperfect cosine response of the collectors, and the results were very similar (less than 15 % difference).

3. *Daily SSA estimation.* The SSA of surface snow is then estimated from the daily albedo spectral dependence. To this end, the snowpack is assumed semi-infinite and uniform. In this case, the diffuse and direct spectral albedos $\alpha_\lambda^{\text{diff}}$ and $\alpha_\lambda^{\text{dir}}$ are related to SSA using the analytical formulation of ~~?~~ Negi et al. (2011) :

$$\alpha_\lambda^{\text{diff}} = \exp\left(-4\sqrt{\frac{2B\gamma_\lambda}{3\rho_{\text{ice}}\text{SSA}(1-g)}}\right) \tag{1}$$

$$\alpha_\lambda^{\text{dir}}(\mu) = \exp\left(-\frac{12}{7}(1+2\mu)\sqrt{\frac{2B\gamma_\lambda}{3\rho_{\text{ice}}\text{SSA}(1-g)}}\right), \tag{2}$$

where $\mu = \cos(\theta)$, θ is the SZA, B and g describe snow single scattering properties and are assumed constant ($B = 1.6$ and $g = 0.86$ after Libois et al., 2014b), $\rho_{\text{ice}} = 917\,\text{kg}\,\text{m}^{-3}$ is the bulk density of ice at $0\,^\circ\text{C}$, and γ_λ is the wavelength-dependent absorption coefficient of ice, taken from Warren and Brandt (2008). SSA is retrieved by minimizing the root mean square deviation between the spectra of daily albedo and the theoretical albedo, accounting for the direct and diffuse components of solar irradiance calculated with SBDART. The comparison is computed in the range 700–1050 nm where the impact of light absorbing impurities is minor (Warren and Wiscombe, 1980) and the sensitivity of snow albedo to snow SSA is high. To ~~deal with the uncertainty on albedo measurement and especially the case it is exceeding 1.0, a~~ account for remaining uncertainties in the albedo spectra, a scaling coefficient A ~~scaling the theoretical albedo~~ is optimized along with SSA, so that the ~~optimized function~~ function to optimize α is actually given by:

$$\alpha_\lambda = A\left[r_\lambda^{\text{diff}}\alpha_\lambda^{\text{diff}}(\text{SSA}) + \left(1 - r_\lambda^{\text{diff}}\right)\alpha_\lambda^{\text{dir}}(\text{SSA})\right], \tag{3}$$

where $r_{\lambda}^{\text{diff}}$ gives the proportion of diffuse light. A is meant to compensate for all the factors affecting in a wavelength-independent way albedo measurements, that were not explicitly corrected by the previous processing steps. This inversion method is somehow similar to estimating SSA from albedo ratios at different wavelengths (Zege et al., 2008).

4. *SSA evolution through the summer.* This procedure is applied every day independently of sky conditions. It is repeated every year from 18 October to 27 February~~every year~~, when SZA at noon remains lower than 67°. Out of this period, we consider that albedo measurements are not accurate enough to retrieve SSA (e.g. Wang and Zender, 2011).

The SSA retrieved with this algorithm roughly corresponds to the SSA of the top ~~2~~$1-2$ cm of the snowpack since the light *e*-folding depth ranges from approximately 5 mm at 1050 nm to 4 cm at 700 nm depending on snow characteristics (Libois et al., 2013). A rigorous forward estimation of the accuracy of the algorithm would require a thorough analysis of several factors including the uncertainty on the collector calibration procedure, the effect of snow anisotropy (Carmagnola et al., 2013) , the shadowing of the surface by the instrument, the potential tilt of the sensor, the validity of the semi-infinite snowpack assumption, etc. Taking into account all these factors and their inter-correlation is beyond the scope of this article and will be addressed in future work. Here the accuracy is estimated using a global approach based on the analysis of the coefficient A obtained during the retrieval. Over the period of observations, it varied in the range 0.98 – 1.03, while ideal measurements would have yielded $A = 1$. The deviation of A from 1, that is -2% to $+3\%$, gives an estimation of the albedo measurement accuracy. Hence we assume that the albedo accuracy is 3%. The corresponding accuracy on SSA estimation is then derived from Eq. 1. For the spectral range of interest and the SSA values encountered at Dome C, the estimated accuracy of the SSA retrieval is better than 25%.

2.1.2 Surface SSA

The SSA of surface snow was also ~~measured manually~~ manually measured during the summer campaigns 2012–2013 and 2013–2014. In 2012–2013, surface SSA was measured using ASSSAP in horizontal position. For this, ASSSAP slides along a 1 m long horizontal rail fixed approximately 5 cm above the surface and measures the snow reflectance at 1310 nm (Fig. 2), from which the surface SSA is estimated with an accuracy of 10–15 % using the algorithm described in Arnaud et al. (2011). Every two days from 22 November 2012 to 16 January 2013 (except from 3 to 6 January), 1 m long horizontal transects of SSA were thus measured without disturbing the snow surface, at 2 fixed locations distant of about 5 m situated 500 m South East of the station (75.10374° S, 123.34093° E). During this summer campaign, the SSA of precipitation particles was also measured at several occasions using ASSSAP sampler. For this, freshly deposited particles were collected on a metallic plate and gathered to fill the sampler. From 27 November 2013 to 29 January 2014, the SSA of snow samples taken from the surface were measured almost every day using ASSSAP sampler. These measurements were taken 100 m further East compared to the previous year (75.10379° S, 123.34484° E), and amounted to a total of 630 snow samples taken randomly in an area of approximately 1000 m^2 over 64 days. As light *e*-folding depth at 1310 nm (~ 2 mm) is smaller than in the range 700–1050 nm, the SSA measured with ASSSAP are not directly comparable to the albedo-derived estimates given that the snow is rarely homogeneous near the surface.

2.1.3 Profiles of SSA

From 23 November 2012 to 16 January 2013, 98 vertical profiles of SSA were measured with ASSSAP from the surface to 10 cm depth at 1 cm resolution (e.g. Carmagnola et al., 2014). The measurements were performed at two different sites. The first one was 600 m West of the main building of Concordia station (75.09971° S, 123.30224° E) and the second one was 500 m South East of the station, just besides the surface SSA measurements. Every day during this period (except from 3 to 6 January), two profiles were measured, one day at the first site, the other day at the second one. All these profiles were taken at different places in undisturbed snow areas with a minimum distance of 5 m between each other. From 25 November 2013 to 25 January 2014, one vertical profile of SSA was measured with ASSSAP every 2 days, amounting to 32 profiles taken in the same area where surface snow samples were taken.

2.2 Satellite observations

The SSA time-series from 1999 to present was estimated from high-frequency microwave radiometers using the approach proposed in Picard et al. (2012). The advantage of observing in the microwave domain is the independence to weather conditions and illumination, which allows to retrieve SSA year-round even during the polar night. To obtain information on the surface snow, we used observations from the Advanced Microwave Sounding Unit (AMSU) constellation that is able to operate up to 150 GHz. Using observations at 150 and 89 GHz makes it possible to estimate the averaged SSA over the top 7 cm approximately. Lower frequencies are more sensitive to snow properties deeper down in the snowpack (Surdyk, 2002; Picard et al., 2009) and present a lower interest for this study.

Following the method of Picard et al. (2012), the DMRT-ML forward microwave emission model (Picard et al., 2013) is used to compute the microwave brightness temperature of an idealized two-layer snowpack. The top layer is 7 cm thick and the bottom one is semi-infinite. The temperature of both layers is set to the 10 day average air temperature taken from ERA-Interim and the density is assumed constant at 320 $\mathrm{kg\,m^{-3}}$ according to the mean surface density reported by Libois et al. (2014a). The SSA in both layers are the unknowns to be estimated. For this, for each 10 day period from 1999 to present, the SSA in both layers is optimized so that the model predictions at 150 and 89 GHz match the satellite observations. To relate the SSA to the grain size metric r required by the DMRT theory, an empirical scaling coefficient is used according to Brucker et al. (2010), such that $\mathrm{SSA} = 3/(\rho_{\mathrm{ice}} \frac{r}{2.8})$.

This method is simple because using 2 observations it considers only 2 unknowns, while the density and layer thickness are probably variable and are known to affect microwave signal as well (even if this effect is of second order compared to the SSA). As a result, this

SSA time-series is not expected to be as accurate as the spectrometry-based approach described in Sect. 2.1.1. The most critical assumption is probably that of constant density. Assuming a density of 300 $\mathrm{kg\,m^{-3}}$ (respectively 350 $\mathrm{kg\,m^{-3}}$) instead of 320 $\mathrm{kg\,m^{-3}}$ yields SSA differences up to +20% (respectively -40%), which gives a broad estimate of 40% for the accuracy of the method. Nevertheless, with 14 years of data it gives a good indication of the inter-annual variability of the seasonal variations of SSA.

2.3 Crocus simulations

2.3.1 Reference simulation (A)

The temporal evolution of snow physical properties at Dome C was computed with the detailed snowpack model Crocus (Vionnet et al., 2012), which simulates the evolution of a one-dimensional multi-layer snowpack in response to meteorological conditions. The ~~snow diagnostic~~ number and thickness of numerical snow layers evolve with time. The snow prognostic properties relevant for this study are snow SSA, snow density, snow temperature and snow sphericity (Carmagnola et al., 2014). Crocus was adapted to the specific meteorological conditions prevailing at Dome C, essentially to handle the very low amount of precipitation, the characteristics of fresh snow, the compaction of snow by the wind during drift events and the rate of metamorphism. In particular, the optimal thicknesses of the 5 topmost layers were set at 2, 3, 5, 5 and 10 mm, to ensure that surface processes are accurately represented. The fresh snow density is fixed at 170 $\mathrm{kg\,m^{-3}}$ and fresh snow SSA is fixed at 100 $\mathrm{m^2\,kg^{-1}}$. All these adaptations are detailed in Libois et al. (2014a). For the present study, a few more modifications were made:

- In Crocus the impact of drift (Vionnet et al., 2012) was originally given in terms of changes in snow dendricity and sphericity (Brun et al., 1992). It was reformulated in terms of SSA ~~by (Carmagnola et al., 2014) (~~using Eq. 5 ~~)~~ of Carmagnola et al. (2014)~~for use with~~ , which is valid for SSA less than 65 $\mathrm{m^2\,kg^{-1}}$. In case of higher SSA as encountered at Dome C, ~~the~~ this formulation leads to SSA decrease during wind drift events. Since such decrease is contradictory to observations in Antarctica

(Kuhn et al., 1977; Grenfell et al., 1994), when the parameterization predicts a decrease of SSA, the latter is actually forced to remain unchanged.

– ~~The simulations are based on~~ SSA decrease is computed from the formulation F06 ~~for~~ of snow metamorphism (Carmagnola et al., 2014) ~~,~~ which is based on ~~an approximation~~ a fit of the semi-empirical microphysical model of SSA decrease rate proposed by Flanner and Zender (2006). ~~Contrary to the original approximation based on a 14fit of this model (Oleson et al., 2010) , here the parameterization used is based on a fit over 100, more adequate to~~ Because of working in the mid-latitude context, the ~~slow~~ fit in Carmagnola et al. (2014) was computed over a period of 14 days, as in Oleson et al. (2010) . Here we use the same approach but extend the period to 100 days to account for the slower metamorphism resulting from the low temperatures prevailing at Dome C. Moreover, below $-50\,^{\circ}\mathrm{C}$, where the parameterization is no more valid, we implemented a scaling of the temperature dependence of snow metamorphism based on Clausius–Clapeyron law for vapor saturation pressure. This latter choice has little impact because at such low temperatures snow metamorphism is anyway negligible.

– The vertical profiles of absorbed solar energy were computed with the physically-based radiative transfer model TARTES (Libois et al., 2013) at 10 nm spectral resolution rather than with the original semi-empirical parameterization implemented in Crocus (Brun et al., 1992). Indeed, TARTES has been fully implemented in Crocus and is used to compute the vertical profile of energy absorption of a multi-layered snowpack based on density and SSA profiles. TARTES also accounts explicitly for snow grain shape, through the asymmetry factor g and the absorption enhancement parameter B. According to Libois et al. (2014b), we chose $B = 1.6$ and $g = 0.86$. TARTES simulates the impact of light absorbing impurities in snow. Here a constant load of black carbon equal to $3\,\mathrm{ng\,g^{-1}}$ was assumed, in agreement with observations (Warren et al., 2006).

- Although in Crocus the roughness length for momentum is usually 10 times larger than that for heat transfer at the air–snow interface, here both were ~~both~~ fixed to 1 mm, as Brun et al. (2011) did in a previous study because this choice produced the best fit between simulated and observed surface temperatures at Dome C.

Crocus was forced by ERA-Interim atmospheric reanalysis for 2 m air temperature and specific humidity, surface pressure, precipitation amount, 10 m wind speed, and downward radiative fluxes. ERA-Interim data were already used by Fréville et al. (2014) to simulate snow surface temperature on the Antarctic Plateau. As detailed in Libois et al. (2014a) , precipitation rate was multiplied by 1.5 to ensure that simulated annual snow accumulation matches observations at Dome C. On the contrary, ERA-Interim wind was found in good agreement with measurements performed on the 40 m high instrumented tower at Dome C (Genthon et al., 2013) .Libois et al. (2014a) also pointed that drift events observed at Dome C could satisfactorily be predicted from ERA-Interim wind time series, further supporting the consistency of wind data. As for air temperature, it does not show any significant bias during the summer from 2000 to 2013 compared to Dome C II automatic weather station (http://amrc.ssec.wis.edu/aws). It does show a positive bias of about 2 K during the winter (Fréville et al., 2014) , but this is not critical for our study because snow metamorphism barely operates in winter.

The snowpack was first initialized with a depth of 12 m ~~, based on~~ using observations of density and SSA at Dome C (Picard et al., 2014). It comprised 25 layers. Crocus was then run 3 times consecutively on the period 2000–2010, which ensured that snow characteristics in the top 2 m of the snowpack were consistently initialized. Then, Crocus was run from 2000 to 2014 and the full state of the snowpack was recorded every 12 h, yielding the reference simulation A ~~-~~that is analysed in the following. The discontinuity in the spinup is not critical here since the analysis focuses priority on the end of the period.

2.3.2 Supplementary simulations (B, C, D, E)

To estimate the sensitivity of Crocus simulations to summer precipitation and air temperature, and to test the hypotheses proposed by Picard et al. (2012) to explain the intensity and inter-annual variability of summer metamorphism at Dome C, 4 additional Crocus simulations were performed and are summarized in Table 1. In simulation B, precipitation is disabled throughout the summer (November–February). In simulation C, 2 m air temperature is increased by 3 K throughout the year to mimic the average warming predicted by the CMIP5 ensemble on the Antarctic Plateau by 2100 for the RCP4.5 scenario (van Oldenborgh et al., 2013). To estimate the strength of the positive feedback between snow albedo and snow metamorphism, in simulation D the snow optical properties are calculated assuming the SSA remains constant, equal to $100\,\mathrm{m^2\,kg^{-1}}$. To isolate the impact of precipitation on summer metamorphism inter-annual variability from that of snow drift which increases snow density and thus decreases snow metamorphism (Flanner and Zender, 2006), in simulation E the 10 m wind speed forcing was taken constant – equal to its mean annual value – throughout the simulation. In this simulation, the density for fresh snow was also increased from the nominal value of 170 to $270\,\mathrm{kg\,m^{-3}}$ to compensate the fact that averaged wind speed is not sufficient to increase snow density through snow drift. This choice ensured that simulated vertical profiles of density remained consistent with the observations.

3 Results

Simulations and measurements show that SSA close to the surface evolves at different time scales. The SSA of the top millimetres is essentially driven by meteorological conditions such as snowfall and drift events (e.g. Grenfell et al., 1994), leading to rapid variations at the daily scale. Deeper, a seasonal decreasing trend is superimposed to these rapid variations. This decrease extends from late October to early February and is highly variable from one year to another. The following sections address these two time-scales.

3.1 ~~Daily~~ Seasonal variations of SSA in the uppermost 2 mm

For each 1 m long horizontal transect taken with ASSSAP in 2012–2013, the average surface SSA in the range 0.25–0.75 m was computed. For that, the measured transect was first divided in 1 cm intervals over which the median SSA value was taken. These medians were then used to compute the average value and standard deviation for the transect, from which the temporal evolution of SSA at the two locations was deduced (Fig. 3a). The main features are the same for both sites, with SSA ranging from about 25 to 90 $m^2\,kg^{-1}$. The periods when precipitation or diamond dust were visually observed at Dome C are also indicated. It highlights that most rapid SSA increases followed precipitation and diamond dust events, as expected because such events bring at the surface snow particles characterized by high SSA as pointed out by Walden et al. (2003) and confirmed by our SSA measurements of precipitation particles that ranged from 90 to 120 $m^2\,kg^{-1}$. SSA generally decreased after fresh snow deposition due to metamorphism (e.g. Taillandier et al., 2007), and it took about 10 days for SSA to drop from approximately 90 to 30 $m^2\,kg^{-1}$. More rapid decreases were observed, like on 14 December 2012, after a strong wind event blew away a thin layer of soft and high SSA snow and let apparent a hard windpacked old snow having low SSA. Erosion rather than snow metamorphism thus explains such rapid changes. The slight continuous increase in SSA observed from 10 to 16 January 2013 is concomitant with the formation of hoar crystals at the surface as reported by Gow (1965) and Champollion et al. (2013). The formation of such crystals may thus contribute to increase snow SSA at the surface (Domine et al., 2009; Gallet et al., 2013).

Figure 3b shows the time series of surface SSA obtained from the snow samples measured during the summer 2013–2014. The SSA of individual samples was in the range 28–185 $m^2\,kg^{-1}$ and the daily median SSA was in the range 35–85 $m^2\,kg^{-1}$. Again, the largest SSA increases occurred after precipitation events. These were followed by periods with SSA decrease. Significant variations also occurred during periods without observed precipitation (e.g. 15–28 December 2013). Snow drift were regularly observed during these periods. The large standard deviation of measurements taken the same day highlights the spatial variability of surface SSA, mainly resulting from snow drift (Libois et al., 2014a). From 12 to 16 January, hoar crystals covered most of the surface and somehow maintained the SSA around 60 $m^2\ kg^{-1}$ despite the absence of precipitation.

These observed SSA variations, corresponding roughly to the top 2 mm of the snowpack, were compared to the reference Crocus simulation A. For this, the average SSA of the top 2 mm was computed from the simulated SSA profiles. The simulated SSA vary in the same range as the measured ones (Fig. 3). In addition, Crocus reproduces relatively well the rapid SSA increases, except when precipitation is not predicted by ERA-Interim (e.g. 1 January 2013 and 25–27 January 2014). The rate of decrease of SSA due to metamorphism in between precipitation events is also correctly simulated for the summer 2012–2013, slightly less for 2013–2014, probably because metamorphism is better captured when an individual snow location is followed throughout the summer, which was the case in 2012–2013, but not in 2013–2014. The effect of soft snow removal by the wind as well as the formation of surface hoar are currently not simulated by Crocus, which explains the discrepancies between model and observations when these processes were observed.

3.2 Seasonal variations of SSA in the uppermost 2 and 10 cm

Beyond these rapid variations of surface SSA, mainly due to snow deposition and transport, the spectral albedo measurements and the vertical profiles show that SSA decreases all along the summer. Figure 4 shows the time series of SSA deduced from spectral albedo measurements, which corresponds approximately to the SSA of the top 2 cm of the snowpack. During the summer 2013–2014, the SSA clearly decreased from 80 to 30 $m^2\ kg^{-1}$ from late October to late January, with most of the decrease occurring before December. This seasonal trend is not fully observed in 2012–2013 because the time series begins on 10 December (date of deployment of the instrument), so that only the rapid variations due to snowfall are visible. Both years are characterized by a large and sudden increase of surface SSA at the end of summer, resulting from a large snowfall (25 February 2013 and 11 February 2014).

The average SSA of the top 2 cm was computed from the SSA profiles simulated by Crocus. With a mean negative bias of $1.2\,\mathrm{m^2\,kg^{-1}}$ and a root mean square difference of $8.1\,\mathrm{m^2\,kg^{-1}}$ over the two summers, the simulated values are in very good agreement with the observations (Fig. 4). Moreover, Crocus successfully simulates the seasonal decrease of SSA observed in 2013–2014, as well as the large increase resulting from the large amounts of fresh snow deposited in February 2013 and 2014. The model also captures the rapid variations of SSA due to precipitation, as already mentioned in Sect. 3.1. ~~Using~~ Since solar irradiance decreases exponentially with depth, the uppermost millimetres of the snowpack contribute more to the albedo than the snow below. As a result, the SSA retrieved from albedo measurements is the result of a convolution of the actual SSA profile by an exponential to a first approximation. To account for this effect, the simulated SSA was also computed using a 2 cm exponential decay (e.g. Mary et al., 2013) rather than a linear average~~to compute SSA~~. This resulted in slightly higher SSA (less than 5 %). ~~Taking the average~~ Likewise, since the choice of 2 cm is to some extent arbitrary, the average was also computed over the topmost 1 ~~or~~ and 4 cm. It resulted in less than 5 % changes, except for the simulated SSA spikes following precipitation that were more marked for 1 cm. Hence the simulated seasonal trend remains consistent with the observations however near-surface SSA is defined.

The summertime decrease of SSA is confirmed by the series of vertical profiles of SSA taken independently with ASSSAP during the same 2 summers (Fig. 5). The average SSA in the top 10 cm decreased from 45 to $28\,\mathrm{m^2\,kg^{-1}}$ in 2012–2013, and from 37 to $29\,\mathrm{m^2\,kg^{-1}}$ in 2013–2014. The summer decrease was thus more significant in 2012–2013 than in 2013–2014, which is reproduced by Crocus (Fig. 5). More precisely, the main difference between both summers is the initial value of SSA. This can be explained by the fact that ERA-Interim precipitation accumulated from March 1st to November 1st was 45% larger in 2012 than in 2013. The SSA decrease observed in 2013–2014 is successfully simulated, even at low temperatures although Crocus was developed for alpine conditions. In the 2012–2013 simulation Crocus underestimates the rapid SSA decrease measured from mid-December. These two independent sets of measurements nevertheless demonstrate that Crocus forced by ERA-Interim is able to simulate the summer variations of SSA close to the surface.

3.3 Inter-annual variability of SSA in the uppermost 10 cm

In situ measurements of SSA down to 10 cm depth are only available for the summers 2012–2013 and 2013–2014. To further understand summer metamorphism at Dome C, the time series of simulated SSA was compared to the SSA estimated from AMSU observations from 2000 to 2014 (Fig. 6). For this, the average SSA of the top 7 cm was computed daily for the simulated snowpack. It was compared to the SSA estimated from the measured snowpack brightness temperatures (Sect. 2.2).

The SSA simulated with Crocus and that deduced from AMSU observations (Fig. 6) are well correlated ($r = 0.70$), which highlights the ability of Crocus to simulate the annual cycle of surface SSA over more than a decade. The SSA values are also in the same range. In particular, the rapid decrease at the end of spring, as well as the slower rate of increase in winter, are similar in the simulation and observations. The rapid increases occurring around 15 February and already observed in Fig. 4 are generally well reproduced and correspond to strong precipitation events (e.g. 2002, 2004, 2011). In contrast, the amplitude of SSA variations is occasionnaly very different. For instance, the SSA decrease during the summers 2001–2002 and 2007–2008 is much larger in the model than suggested by satellite observations. Moreover, the simulated SSA increased almost as usual in winters 2007 and 2010 while AMSU data suggest that this increase was much less than usual. AMSU observations nevertheless confirm that summer metamorphism was more intense in 2012–2013 than in 2013–2014, as noted from in situ measurements (Fig. 5).

4 Discussion

A version of Crocus adapted to the meteorological conditions of the Antarctic Plateau was used to simulate the temporal variability of snow SSA close to the surface at Dome C, in order to identify the physical processes responsible for summertime SSA variations. In general, a satisfactory agreement was obtained with regards to in situ measurements and remote sensing observations of snow SSA, even though some discrepancies remained between model and observations.

4.1 Metamorphism, snowfall and wind driven SSA variations

During the winter period at Dome C, which extends approximately from late February to mid October, snowfalls deposit onto the surface fresh snow whose detailed characteristics generally depend on weather conditions, but whose SSA is invariably high. Snow metamorphism is very limited during this period due to the prevailing extremely low temperatures. As a consequence, at the end of winter, snow properties in the layer accumulated during this period ($\sim 6\,\mathrm{cm}$ according to Libois et al., 2014a) mainly reflect the properties of winter precipitation. In late October, as solar radiation becomes stronger and air temperature increases, snow metamorphism starts, conducting to an approximate three-fold decrease of SSA by mid February. The time of initiation of summer metamorphism in Crocus simulation is very consistent with the observations, as well as the date when minimum SSA is reached. Conversely, the amplitude of the SSA decrease is more contrasted between observations and model. Supplementary simulations were thus performed to investigate what drives the amplitude of summertime metamorphism in Crocus.

The results of simulation B, where summer precipitation was inhibited, imply that snow metamorphism only weakly depends on the total amount of precipitation during summer (Fig. 7), probably less than proposed by Picard et al. (2012). Indeed, the minimum average SSA for the topmost $7\,\mathrm{cm}$ at the end of summer simulated by Crocus, that is used to quantify the intensity of metamorphism, is on average only $6\,\%$ lower in simulation B than in simulation A. This metamorphism indicator is chosen instead of using the decrease in SSA, because the latter is highly dependent on the SSA value at the end of the previous winter, a quantity that is likely to be erroneous in Crocus simulations. High SSA snow actually deposited during snowfall increases the overall SSA, but also reduces snow metamorphism by inhibiting the positive feedback as proposed in Picard et al. (2012). To quantify the importance of this feedback, we run Crocus with a constant SSA ($100\,\mathrm{m^2\,kg^{-1}}$) as input of the radiative transfer calculations performed with TARTES to deactivate the link between snow SSA variations and albedo, all other things being equal (simulation D). This resulted in less intense summer metamorphism (Fig. 7), with the SSA at the end of summer 15 % higher than in simulation A, which is consistent with simulations using a simpler model than Crocus but with similar optical scheme, as noticed by Picard et al. (2012). This suggests that using a fine representation of snow optical properties which accounts for snow properties evolution is essential to correctly simulate SSA evolution at Dome C. The results of simulation C, where 2 m air temperature was increased by 3 K year-long, show a 12 % lower SSA at the end of summer with respect to the reference simulation. This shows that the direct effect of atmospheric warming on dry snow metamorphism is likely to remain moderate over the XXIst century. This also highlights the primordial role of feedback loops. Overall, the sensitivity of simulated summer SSA decrease to air temperature and precipitation is relatively weak. This probably explains why the SSA decrease is less variable from one year to another ~~than~~ in Crocus than in the observations (Fig. 7).

Although the impact of precipitation seems moderate in Crocus simulations at the seasonal scale, snowfall occurrence and amount drive ~~SSA variations at shorter time scales, which is consistent~~ Crocus-simulated SSA variations in the top 2 mm, consistently with field observations. While the deeper layers show a seasonal SSA evolution, the surface layer mostly reflects day-to-day variations of weather conditions. To simulate the evolution of the snowpack at Dome C, it is thus critical to know precipitation very precisely, a quantity that is difficult to obtain from reanalyses in Antarctica (Bromwich et al., 2011). In practice, ERA-Interim reanalysis sometimes misses precipitation events at Dome C (Fig. 3), which explains most differences between observations and simulations. More generally, shortcomings in parameterizing the surface boundary layer on the Antarctic Plateau sometimes conduct to poorly simulated air temperature and wind profiles (Genthon et al., 2010), which probably participates to these differences as well. Indeed, besides precipitation, surface snow at Dome C is also largely shaped by snow drift, which redistributes snow and controls density and SSA in the topmost centimetres (Gallée et al., 2001; Albert et al., 2004). Snow drift also generates spatial variability of snow properties close to the surface because it can accumulate fresh snow at some locations and make apparent older snow at other locations through erosion (Libois et al., 2014a), which is illustrated by the large standard deviation of surface SSA measurements in Fig. 3b. This spatial heterogeneity is difficult to simulate and makes complicated the comparison between punctual observations and simulations (Groot Zwaaftink et al., 2013). In particular, the horizontal transects of SSA used in this study are representative of approximately 1 cm × 1 m, which is not sufficient to capture this spatial heterogeneity, especially in the topmost millimetres of the snowpack. This might explain the discrepancies with Crocus simulation for the rapid variations of surface SSA. Conversely, the spectral albedo measurements cover an area ~~of approximately 3 m × 3 m~~ with radius of approximately 6 m (Schwerdtfeger, 1976) and probe deeper into the snowpack~~, which is~~. Hence they are more likely to be representative of ~~surface snow~~ the average snow SSA in the topmost 2 cm at Dome C, even though larger-scale spatial variability exists (e.g. Picard et al., 2014). This probably explains the success of Crocus to simulate the SSA variations derived from spectral albedo measurements (Fig. 4).

Despite a few deviations from the observations, Crocus captured well the variations of SSA in response to meteorological conditions and metamorphism at Dome C. Since metamorphism strongly depends on the temperature profile close to the surface, this suggests that Crocus successfully resolves the energy budget of the snowpack close to the surface, as already pointed out by Brun et al. (2011) and Fréville et al. (2014). It also proves that the metamorphism parameterization of Flanner and Zender (2006) is appropriate to study snow on the Antarctic Plateau, although this parameterization cannot be strictly assessed in complex conditions as encountered at Dome C. This is promising for larger scale studies over the Antarctic Plateau, and puts Crocus as an appropriate tool to investigate the spatial pattern of SSA over the Antarctic continent (Scambos et al., 2007), which probably results from the combined effects of precipitation, snow drift, and metamorphism.

4.2 Inter-annual variability of summer metamorphism

The fact that Crocus poorly simulates the inter-annual variability of SSA summer decrease, while it proved efficient to simulate the seasonal variations, is more puzzling. Actually, the apparent intensity of the metamorphism depends both on the SSA value at the end of winter, and on the rate of SSA decrease during summer, which are driven by different processes.

The differences between simulated and observed SSA at the end of winter (Fig. 6) can be attributed either to inaccuracies in ERA-Interim precipitation or to the simple treatment of fresh snow characteristics in Crocus. For instance, the large and sudden increase of SSA simulated in January 2007 (Fig. 6) results from a strong deposition event forecasted by ERA-Interim reanalysis, probably stronger than the actual event. This overestimation might result from the large horizontal scale of the reanalysis (~ 50 km). As for fresh snow SSA in Crocus, it was assumed constant (100 $m^2\,kg^{-1}$), based on our observations of summer precipitation at Dome C. In winter, due to colder conditions, the SSA of fresh snow might be higher, though, as suggested by the observations of Walden et al. (2003) who measured SSA of diamond dust up to 300 $m^2\,kg^{-1}$. This highlights the need to extend the existing parameterizations of fresh snow properties developed in the Alps to polar regions.

As to the summer decrease in SSA, Picard et al. (2012) found a strong correlation between AMSU estimated metamorphism amplitude and the total amount of summer precipitation predicted by ERA-Interim (in their study, summer refers to the period 1 December–15 January). Figure 8 shows the minimum SSA (topmost 7 cm) over this summer period in terms of accumulated precipitation for the reference simulation A and for simulation E, where wind remains weak throughout the year and drift events are thus inhibited. For the reference simulation, there is no significant correlation, which is contradictory to Picard et al. (2012) and to the observed influence of snowfall on the rapid variations of SSA in the present paper. On the contrary, for simulation E the correlation is significant ($r = 0.81$), in agreement with AMSU observations. This suggests that the impact of wind on snow SSA may be too strong in simulation A. In Crocus, snow drift increases surface SSA, but also increases density through compaction, which decreases metamorphism rate (Flanner and Zender, 2006). The apparent deficiency of the reference simulation can be attributed either to an inappropriate parameterization of snow compaction and SSA increase by the wind, or to an over-sensitivity of snow metamophism in Crocus which may be incorrect. This makes wind a major driver of snow metamorphism in the reference simulation, and highlights the need to further understand the impact of drift events on surface snow density.

5 Conclusions

Crocus simulations suited to the meteorological conditions of the Antarctic Plateau were compared to in situ and satellite derived measurements of snow SSA at Dome C, in order to identify the processes controlling SSA evolution on the Antarctic Plateau. The observations show rapid variations of SSA close to the surface, mainly due to precipitation and snow drift. They also confirm the existence of a seasonal cycle of SSA in the topmost 10 cm of the snowpack, characterised by a two-to-three-fold decrease of SSA in summer and a slower, continuous increase in winter due to accumulation of precipitation crystals with high SSA. These variations of SSA are successfully simulated by Crocus, provided the meteorological forcing is adequate. In particular, the intensity of the summer metamorphism and the date of its initiation agree well with the observations. However, the inter-annual variability of the summer decrease in SSA is not well captured, probably because the parameterization of the effect of snow drift on snow SSA is too strong in the model. This study demonstrates that Crocus can capture the main features of snow metamorphism in the conditions of Dome C for which it was not originally developed, which is promising for extended studies of surface snow SSA and evolution at the scale of the Antarctic Plateau or whole continent. Nevertheless, SSA is very dependent on the occurrence and intensity of precipitation events, which are known to be difficult to predict by reanalysis, highlighting the need to further improve the characterisation of precipitation in this high and extremely dry region. Other physical processes not yet simulated by Crocus should also be regarded as potential progress for simulating snow properties on the Antarctic Plateau, such as the formation of hoar crystals, and the mixing of the topmost layers of the snowpack due to snow drift.

Data availability

The data used in this study are available upon request from the authors (ghislain.picard@ujf-grenoble.fr). The Crocus simulations were performed with SURFEX v7.3, adapted to Antarctic conditions. The version of Crocus incorporating the specific developments described in this article has not yet been officially released but is available upon request to crocus@meteo.fr.

Author contributions. Q. Libois, G. Picard, L. Arnaud and E. Lefebvre participated to in situ measurements at Dome C. Q. Libois developed the Antarctic parameterizations of Crocus and performed the corresponding simulations. G. Picard performed the DMRT-ML simulations. Q. Libois and G. Picard analysed satellite and field data. M. Dumont contributed to the parameterization of incident solar radiation in TARTES. M. Lafaysse implemented the model TARTES in Crocus. S. Morin helped with Crocus simulations and implemented the Antarctic parameterizations in the code. Q. Libois and G. Picard prepared the manuscript with contributions from M. Dumont, M. Lafaysse and S. Morin.

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Table 1. Crocus simulations performed for this study.

Simulation	Characteristics
A	Reference
B	Same as A, precipitation set to 0 from November to February
C	Same as A, 2 m air temperature increased by 3 K
D	Same as A, SSA kept constant at 100 $m^2 kg^{-1}$ for calculations of optical properties
E	Same as A, wind speed in the forcing is constant, equal to the average ERA-Interim wind speed over the simulation period (5 $m s^{-1}$), fresh snow density is set to 270 $kg m^{-3}$

Figure 1. Spectral albedo measurement at Dome C (picture taken on 12 January 2014, 11:00 local time). The vertical ~~mat~~ mast is approximately ~~2.2~~2 m high.

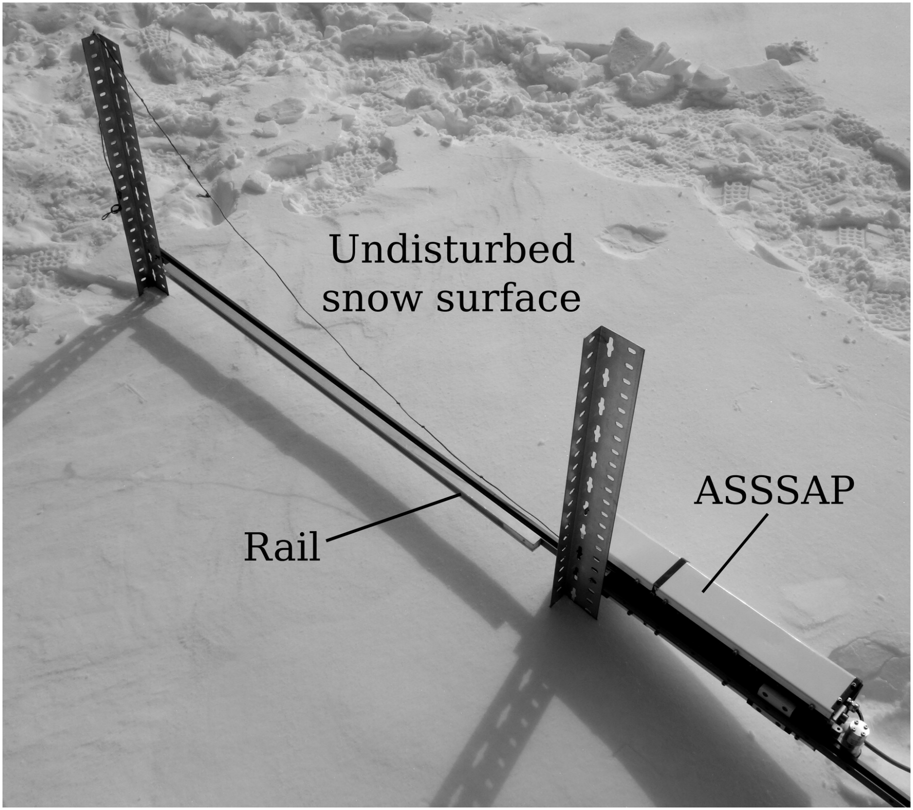

Figure 2. Experimental setup for transect measurements of SSA using ASSSAP in horizontal position. The distance between both vertical stakes is 1 m. ASSSAP slides along the horizontal rail and records the SSA of the surface beneath every 10 ms. For the measurements, the setup was covered by a dark tarpaulin to avoid the supersaturation of ASSSAP photodiodes.

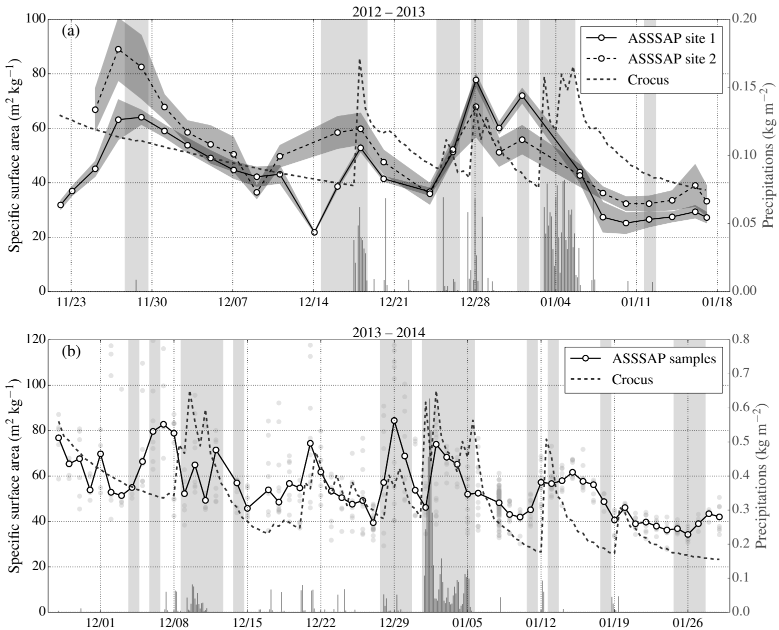

Figure 3. (a) Evolution of surface SSA during the summer 2012–2013 deduced from the 1 m long horizontal transects taken with ASSSAP at 2 distinct locations (ASSSAP 1 and 2), and SSA of the top 2 mm simulated with Crocus. The points show the mean value of each transect and the standard deviation is indicated by the ~~hatched~~ shaded area. **(b)** Evolution of surface SSA during the summer 2013–2014 deduced from the snow samples measured with ASSSAP, and SSA of the top 2 mm simulated with Crocus. The ~~clear dots~~ grey circles indicate single measurements and the ~~dark line highlights~~ white circles highlight the median value for each day. In (**a** and **b**), the shaded bands indicate the periods of observed snowfall or diamond dust at Dome C. The amount of precipitation predicted by ERA-Interim is also shown (right y-axis, dark grey columns).

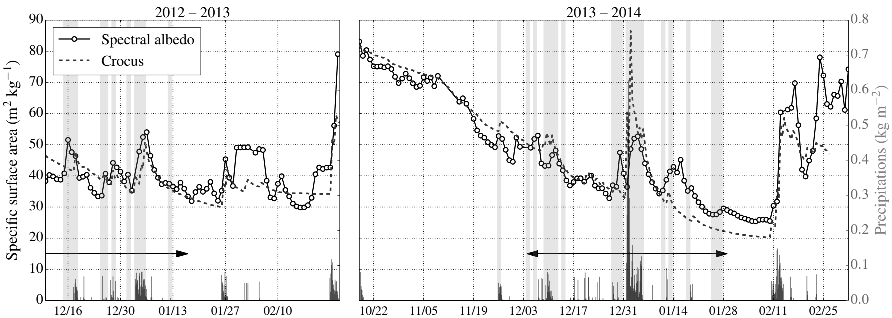

Figure 4. Variations of SSA close to the surface deduced from the spectral albedo measurements, and average SSA of the top 2 cm of the snowpack simulated with Crocus, for the summers 2012–2013 and 2013–2014. The shaded bands indicate the periods of observed snowfall at Dome C. The amount of precipitation predicted by ERA-Interim is also shown (right y-axis, dark grey columns). The horizontal arrows highlight the periods of measurements shown in Fig. 3.

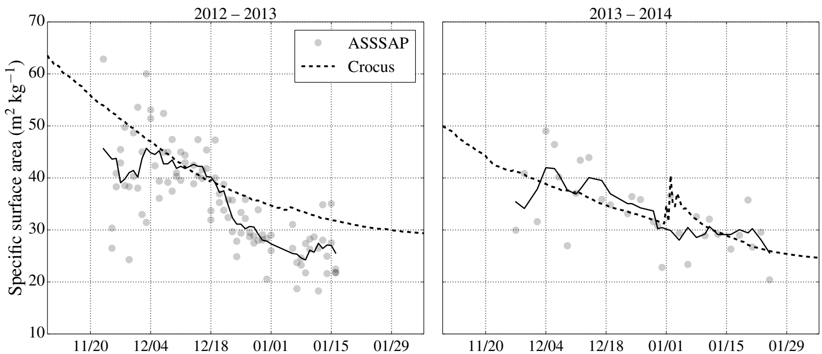

Figure 5. Evolution of SSA averaged over the top 10 cm, deduced from vertical profiles of SSA taken with ASSSAP for the summers 2012–2013 and 2013–2014, and simulated with Crocus. The integrated value from each profile is indicated by a circle. The continuous lines correspond to the 4 day moving averages of the measurements.

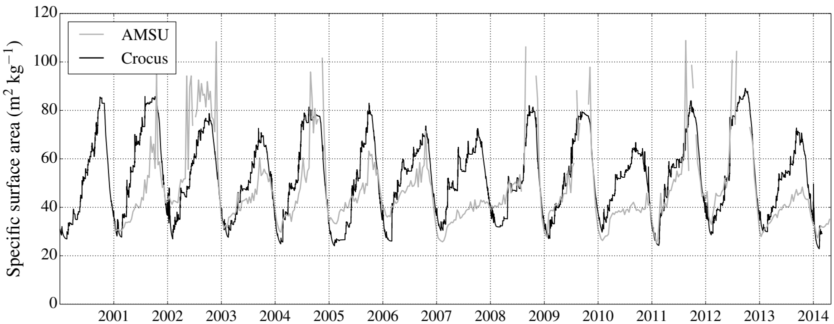

Figure 6. Comparison of SSA evolution deduced from AMSU brightness temperatures at 89 and 150 GHz, and simulated with Crocus (top 7 cm).

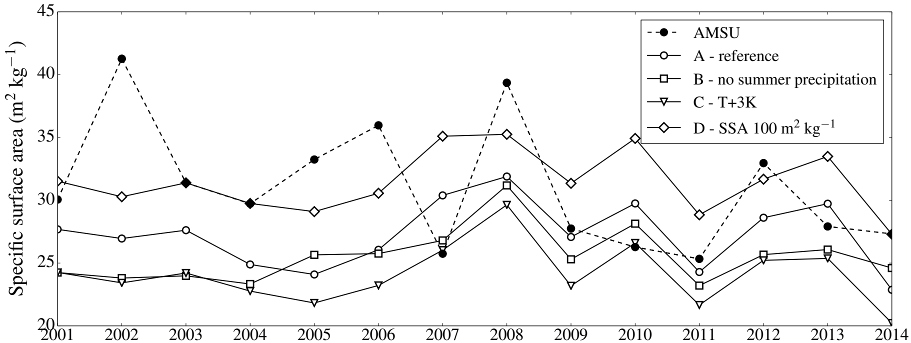

Figure 7. Minimum SSA (top 7 cm) at the end of summer for AMSU estimations and Crocus simulations A, B, C and D, from 2001 to 2014.

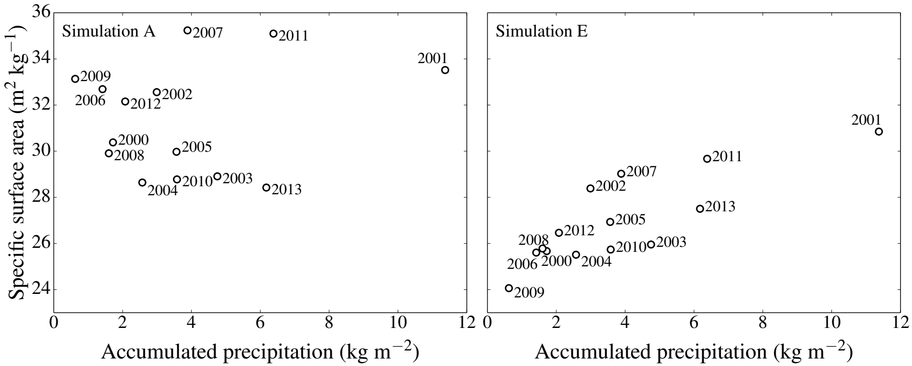

Figure 8. Minimum SSA of the topmost 7 cm of the snowpack simulated by Crocus for each summer (1 December–15 January), vs. accumulated precipitation during this period, for simulations A and E.